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FALLSPEEDS AND VERTICAL AIR MOTIONS IN STRATIFORM RAIN DERIVED FROM ER-2 DOPPLER RADAR OBSERVATIONS

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1. INTRODUCTION

The Tropical Rain Measuring Mission (TRMM) conducted several intensive field validation campaigns for improved understanding of Tropical precipitation systems (Zipser et al., this proceeding). Two of the campaigns (TEFLUN in Florida and Texas, and LBA in Brazil) utilized the NASA ER-2 high-altitude (20 km) remote sensing aircraft instrumented with the ER-2 Doppler Radar (EDOP), the University of North Dakota Citation microphysics aircraft, and the NCAR S-POL polarization radar. This paper focuses on EDOP-derived fallspeeds and vertical velocities in the rain regions of two stratiform cases (5 September 1998 along the east coast of Florida, and 17 February 1999 in Amazonia in Brazil). These cases were sampled in situ microphysically by the Citation and reported elsewhere in this meeting; the main emphasis of this paper will be on the airborne radar measurements and inferences from them.

2. BACKGROUND AND APPROACH

EDOP is an X-band (9.6 GHz) nadir pointing Doppler radar. EDOP samples Doppler velocity, v_D, and reflectivity factor, Z, with a gate spacing of 37.5 m (vertical resolution) and at 0.5 s intervals (100 m along-track sampling frequency). The antenna beamwidth is about 3°, a footprint of about 1.2 km at the surface. Calibration accuracy of EDOP is about 1 dB. EDOP provides similar measurements to ground-based vertically pointing radars except that extensive regions can be covered in short times.

Doppler velocities v_D from a nadir-pointing airborne radar represent the sum of the mean reflectivity-weighted hydrometeor motion, the vertical air motion, and the aircraft vertical motion. Atlas et al. (1972) derived relations between Doppler velocity, reflectivity, and rain rate assuming an exponential size distribution for rain. Ulbrich (1994) expanded on this work assuming a Gamma size distribution. Retrieval of information on raindrop size distributions with the above method requires that air and aircraft motions be removed from the observed Doppler velocities.

The radar reflectivity factor Z in $mm^6 m^{-3}$ is given by:

$$Z = \int_{D_{min}}^{D_{max}} N(D) D^6 dD \qquad (1)$$

where D (cm) is particle equivalent spherical diameter, $N(D)$ in $m^{-3} cm^{-1}$ is number of raindrops per unit volume per unit size interval, D_o is the median particle diameter. For a Gamma size distribution, $N(D) = N_o D^{\mu} e^{-\Lambda D}$, where μ is the shape parameter, and $\alpha = \Lambda D_o = 3.67 + \mu + 10^{-0.3(9+\mu)}$. $N_o(\mu) \approx 6.4 \times 10^{4} \exp(3.2\mu)$) has been shown observationally and theoretically by Ulbrich (1983). The integration limits D_{min} and D_{max} are assumed $0 \rightarrow \infty$, though several papers have dealt with truncated limits. The reflectivity-weighted Doppler velocity v_Z (ms^{-1}) is given by:

$$v_Z = \frac{\int_o^{\infty} v(D) N(D) D^6 dD}{\int_o^{\infty} N(D) D^6 dD} \qquad (2)$$

where $v(D)$ is the drop fallspeed. Furthermore, Ulbrich (1994) has derived the reflectivity-weighted Doppler velocity v_Z as a function of μ by assuming the Gunn-Kinzer fallspeed relation at ground level (1013 mb), i.e., $v(D) = 9.65 - 10.3\, e^{-6D}$. Assuming that vertical air motions are neglible, Eq. 2 can be given as a function of μ and Z:

$$v_Z(\mu) = 9.65 - 10.3 \left\{ 1 + 6 \left[\frac{Z}{N_o(\mu) 10^6 \Gamma(7+\mu)} \right]^{1/(7+\mu)} \right\}^{-(7-\mu)} \qquad (3)$$

This relation applies to ground level (1013 mb); v_Z at other altitudes is obtained by multiplying it by $[\rho_o/\rho]^{0.44}$ where ρ and ρ_o are the air density at the surface and measurement height, respectively. Eq. 3 has a major advantage over power law relations used extensively in the literature: v_Z is asymptotic with increasing reflectivity rather than increasing monotonically with Z.

Figure 1 shows the dependence of v_Z on μ using Eq. 3; a commonly used power law relation is also

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plotted. At 40 dBZ, the variation of μ from -2 to 6. results a fairly substantial 4 ms^{-1} variation in v_Z. Figure 2 shows the Gamma size distribution corresponding to v_Z = 6, 8 ms^{-1} and μ =-2 to 4. The resulting range of generated curves are typical of reflectivities observed in EDOP data sets. Larger v_Z for a given μ implies larger D_o and a larger Z.

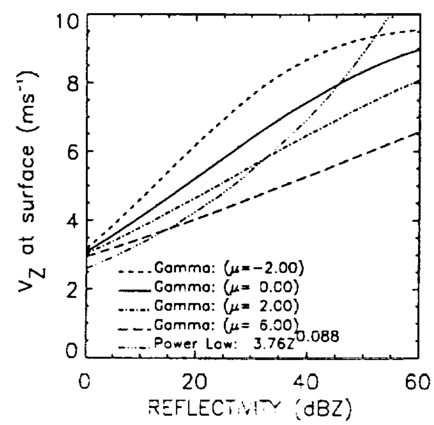


Figure 1. v_Z - Z relation as a function of μ.

3. EXAMPLES.

If one has large sample of v_D and Z measurements from the same region, then μ can be obtained by fitting Eq. 3. This requires the assumption that stratiform regions are fairly extensive. However, Eq. 3 must be modified with an additive constant C since widespread vertical motions can be present in the stratiform region that will bias fits to Eq. 3. Smaller scale vertical motions can also be present in the stratiform but are ignored in the following. Thus, the $v_D - Z$ data from EDOP is fitted with the following:

$$\left(\frac{\rho}{\rho_o}\right)^{0.44} v_D = C + v_Z(\mu), \qquad (4)$$

where v_Z is obtained from Eq. 3, C is a constant which consists of the mean mesoscale vertical motion, positive (negative) C indicates downdrafts (updrafts), and the altitude dependence of the measurements are removed by the density term on the left side. This fitting method provides a global estimate of μ and C within some defined height interval, not a point-by-point estimate of the values. Besides the inhomogeniety

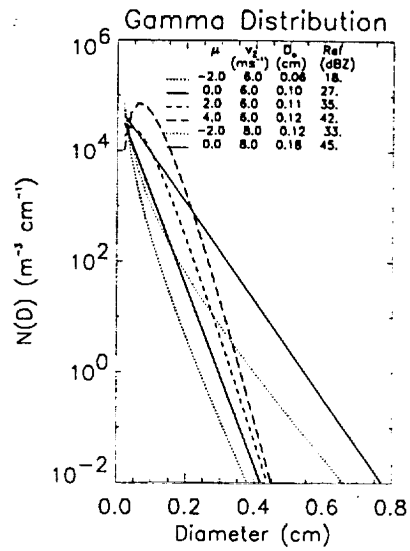
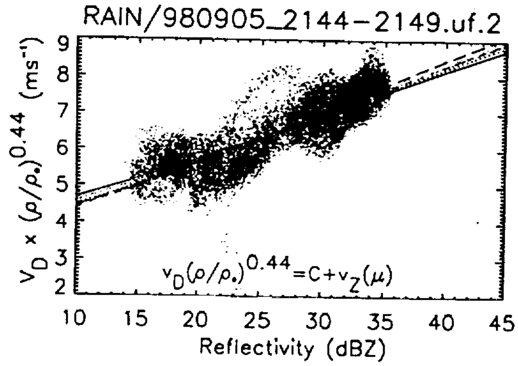


Figure 2. Gamma distribution for two values of v_Z, assuming various values of the shape parameter μ.

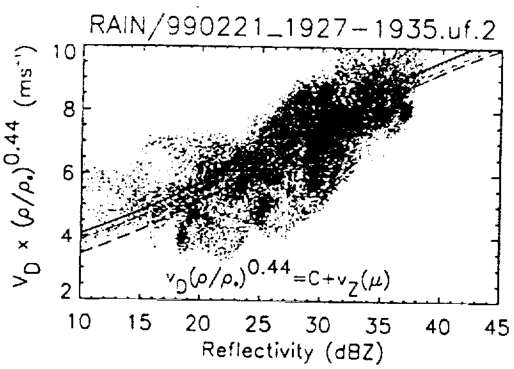
of the size distribution over the analyzed region, a major source of error in this approach is the presence of small scale vertical air motions. In addition, small scale aircraft motions are not always fully removed from v_D resulting in an error in the calculation.

Figures 3 and 4 show an examples of applying Eq. 4 to stratiform regions in Florida during TEFLUN-B and Brazil during TRMM-LBA. The rain region between the surface and 400 m below the freezing level, is divided into approximately 500 m intervals and then fitted using a non-linear curve fitting routine. The vertical axis corresponds to observed Doppler velocities adjusted to the surface. The number of points used in Fig. 3 and Fig. 4 per layer are 6552 and 12000, respectively. It is noted that points above (below) the fitted curve are due to smaller scale downdrafts (updrafts) assuming aircraft motions have been removed. The scatter of the points around the fits is likely due to inadequate aircraft motion removal. Nevertheless, the fits are statistically significant, with an obvious increase in v_D with Z. The fitted curves in Fig. 3 for stratiform in Florida indicate that μ ranges from −1.4 at the top layer (2.74-3.19 km) to -0.4 at the bottom layer (0.3-0.75 km). Figure 4 indicates lower μ values (-3) which suggest a narrower size distribution than for the Florida flight line in Fig. 3. The calculated C ranges from about 1.5-2.0 ms^{-1} in the rain layers of Figs. 3 and 4 which suggests the presence of a mesoscale downdraft in both cases. Vertical reflectivity images for both cases suggest evaporation is possibly occurring for both cases which would account for the subsidence.



Bottom (km)	Top (km)	Npts	C (ms^{-1})	μ	N_o (m^{-3}cm$^{-1-\mu}$)
0.30	0.75	6552	1.7	−0.4	1.6E+04
0.79	1.24	6552	1.6	−1.0	3.0E+03
1.28	1.73	6552	1.6	−1.2	1.2E+03
1.76	2.21	6552	1.7	−0.9	3.3E+03
[illegible]	[illegible]	[illegible]	[illegible]	−1.0	[illegible]
2.74	3.19	6552	1.7	−1.4	6.5E+02

Figure 3. Observed measurements from stratiform on 5 September 1998 fitted with Eq. 4.



Bottom (km)	Top (km)	Npts	C (ms^{-1})	μ	N_o (m^{-3}cm$^{-1-\mu}$)
0.30	0.75	11812	2.0	−3.3	1.9E+00
0.79	1.24	11935	1.9	−3.4	1.2E+00
1.28	1.73	12119	1.8	−3.2	2.2E+00
1.76	2.21	12181	1.7	−3.1	2.7E+00
[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]
2.74	3.19	12181	1.5	−3.5	9.0E−01

Figure 4. Same as Fig. 3 except for 21 February 1999 fitted with Eq. 4.

4. CONCLUSIONS.

A "global" fitting procedure has been presented to obtain layer values of μ in the Gamma distribution, and mean vertical air motions in stratiform rain regions. This approach facilitates a better understanding of the observational errors (aircraft motions) and the inhomogenieties in air motions and size distributions. The calculated μ for the two presented is less than 0, thus implying a narrow distribution of raindrops. This is in contrast to $\mu>0$ reported in the literature. The radar measurements and the approach presented will be explored further using insitu microphysics data and ground-based polarization radar data collected along with the EDOP measurements.

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